



Mineralization of the Hansonburg Mining District, Bingham, New Mexico

John Rakovan and Frederick Partey

2009, pp. 387-398. <https://doi.org/10.56577/FFC-60.387>

in:

Geology of the Chupadera Mesa, Lueth, Virgil; Lucas, Spencer G.; Chamberlin, Richard M.; [eds.], New Mexico Geological Society 60th Annual Fall Field Conference Guidebook, 438 p. <https://doi.org/10.56577/FFC-60>

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MINERALIZATION OF THE HANSONBURG MINING DISTRICT, BINGHAM, NEW MEXICO

JOHN RAKOVAN¹ AND FREDERICK PARTEY²

¹Department of Geology Miami University, Oxford, Ohio 45056 rakovajf@muohio.edu

²New Mexico Bureau of Geology and Mineral Resources, Socorro, NM, 87801

ABSTRACT—The Hansonburg Mining District, Socorro County, New Mexico is the largest of the hydrothermal Rio Grande Rift fluorite-barite-galena deposits that are found within and along the southern half of the rift. This paper reviews key aspects of the history, structure, geochemistry and mineralogy of the district that have been extensively studied for over half a century. Recent work has elucidated the trace element chemistry of fluorite, geochronology and stable isotope geochemistry, and the origin of fluorine and other key constituents in hypogene minerals of the of the district, as well as other related Rio Grande Rift - type deposits.

INTRODUCTION

The Hansonburg Mining District, as it is most commonly recognized today (see historical notes below), is located in south-central New Mexico, approximately 28 miles east of San Antonio along U.S Highway 380, and approximately 5 miles south of Bingham, Socorro County (Fig. 1). The district is located along the eastern edge of the Rio Grande Rift system, which is characterized by north-south trending normal faults related to regional extension (Chapin, 1979; McMillan, 1998) that began in the middle Cenozoic (36 m.y. ago; McMillan et al., 2000). The mines themselves lie along the western flank of the Sierra Oscura Mountains and trend north-south for about 2.5 miles (Fig. 2).

The Hansonburg Mining District is one of about 30 barite-fluorite-galena deposits found throughout the Rio Grande Rift in southern New Mexico. In the past, these were classified as Mississippi Valley type deposits (Roeder et al., 1968; Allmedinger, 1975; Putnam et al., 1983; Norman et al., 1985), but because of several significant differences, including the abundance of fluorite several studies have suggested that they be classified as a new type of deposit referred to as Rio Grande Rift barite-fluorite-galena deposits (McLemore and Lueth, 1996; Sutphin, 1997; McLemore et al., 1998; Parthey et al., 2009). The most common primary (hypogene) minerals in the Hansonburg Mining Dis-

trict are fluorite, barite, galena, and quartz, with minor amounts of other sulfide minerals (sphalerite, chalcopyrite and pyrite).

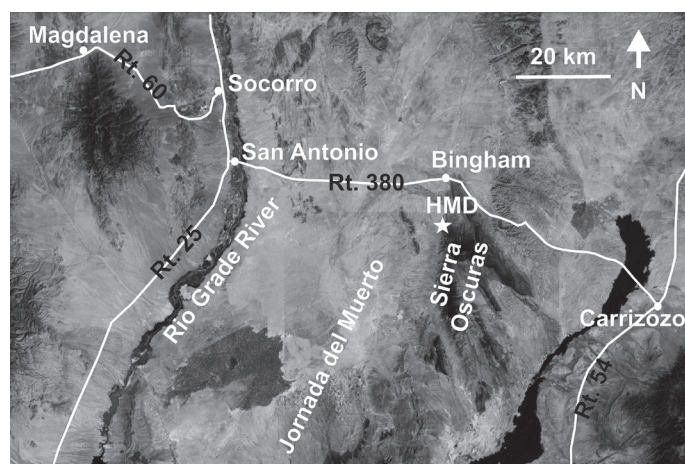


FIGURE 1. Satellite image of South Central New Mexico and the Hansonburg Mining District (HMD). Image © DigitalGlobe © 2009 Tele Atlas Image NMRGIS.

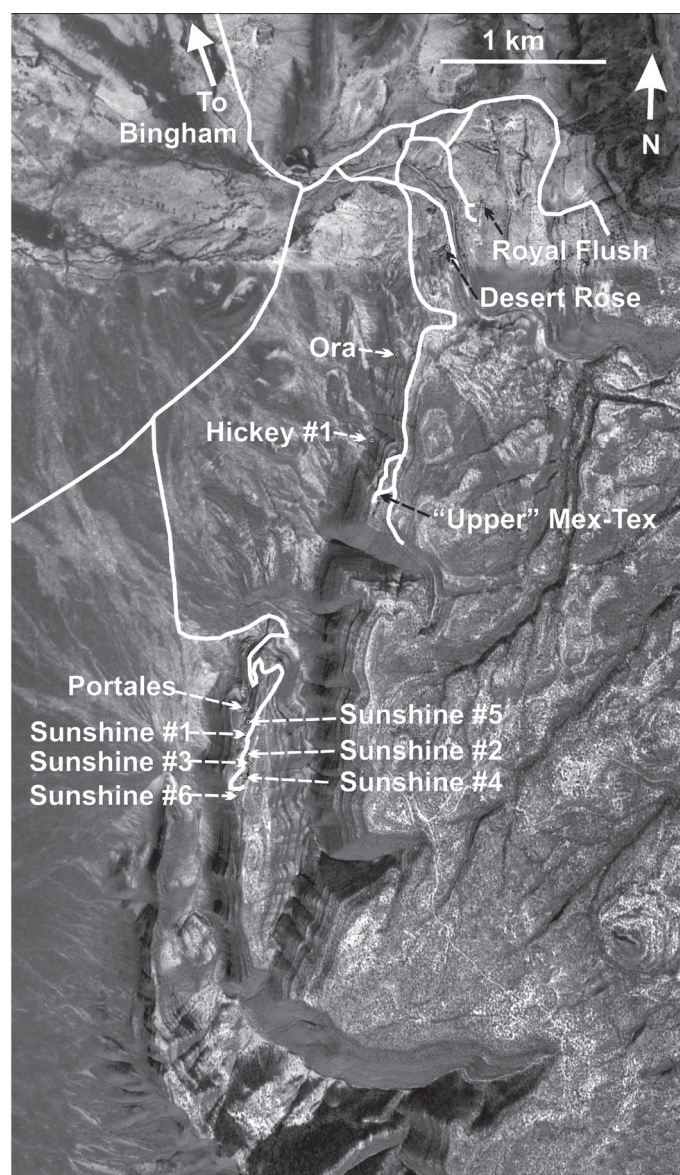


FIGURE 2. Satellite image of the Hansonburg Mining District with mine locations and pertinent roads. Image © DigitalGlobe © 2009 Tele Atlas Image NMRGIS.

Secondary mineralization, although volumetrically very small, includes many interesting Pb, Cu and Zn carbonates and sulphates. The mineralization typically fills or lines dissolution cavities or is found in large laterally extensive bands, along bedding planes. Although the Hansonburg Mining District is rich in barite-fluorite-galena ores it has seen relatively little metal production throughout its history because of metallurgical problems in separating the mixed ores, and lack of water for mining operations (Norman et al., 1985). Its most significant production was in barite. Since 1987 all of the mines in the district have been under claim by individuals who have developed them for mineral specimens.

This paper reviews key aspects of the history, structure, geochemistry and mineralogy of the Hansonburg Mining District. Recent studies have elucidated the trace element chemistry of fluorite (Hill, 2000; Bosze and Rakovan, 2002; Bosze et al., in press), the geochronology and stable isotope geochemistry (Lueth et al., 1999; Hill, 2000; Lueth et al., 2005), and the origin of fluorine as well as other key constituents in the ore minerals of the deposit and other related Rio Grande Rift Deposits (Lueth et al., 2005; Partey et al., 2009). Previous reviews of the district can be found in Kottlowski (1953) Kottlowski and Steensma (1979), Putnam et al. (1983), Norman et al. (1985), Sharp (1988) and Taggart et al. (1989).

HISTORY

Today the Hansonburg Mining District, Bingham, NM is almost solely associated with the barite-fluorite-galena mineralization found along the Western slopes of the Sierra Oscura Mountains, and it is that deposit which we constrain this review article to. However, the history of the district includes, and the name was actually first applied to, a copper deposit in the Jornada del Muerto between the Rio Grande River and the Sierra Oscuras. The formation and mineralogy of these two deposits are completely distinct, and a detailed history of the name (HMD) and the associated copper deposit is given by Eveleth and Lueth (2009).

There are numerous individual mines (mostly single adits) within the Hansonburg District. Some of these have changed names over time and some are known by different names. Taggart et al. (1989) combined the mines of the district into four groups that are geographically separated. From north to south (Fig. 2) these include the Royal Flush group with the Royal Flush Mine (North and South adits) and the Desert Rose Mine (a.k.a. Downey Stope, Mountain Canyon Mine, Ace High Mine); the Lower Mex-Tex group with the Ora Mine, Caliche, and Hickey #1 Mine; and the Upper Mex-Tex Mine group; including the Snake Pit, the Byrd Tunnel and pit, Upper Mex-Tex Mine (a.k.a. Julian-Malachite Mines; Kottlowski 1953), and a small unnamed adit at the south end of the group (DeMark and Massis, 1999). About 1.2 km south of the Mex-Tex, on a separate fault block that forms a distinct N-S trending, elongate hill (sometimes referred to as Hansonburg Hill; Norman et al., 1985), lies the Blanchard Claims or Blanchard Mine group (Fig. 3). This includes the Portales (a.k.a. MacCarthy Lead Mine, Hansonburg Lead Mine, Blanchard glory hole, or Blanchard Mine), Sunshine #5 adit,

Sunshine #1 adit, Sunshine #2 adit, Sunshine #3 adit, Sunshine #4 adit, and the Sunshine #6 adit (Kottlowski, 1953; DeMark, 2003). At the southern most extent of the district is the Barrett workings. Throughout the district there are numerous surface cuts, trenches and prospects that are unnamed here.

Although mineral exploration in the barite-fluorite-galena deposit of the district predates Spanish and Anglo developments, the first major mining operation started in 1916 by Western Mineral Products Company, who also constructed a 50 ton dry mill on the site to extract galena. They started several excavations including an adit that would later become the Portales mine of the Blanchard Family. The name Blanchard is inextricably associated with the Hansonburg Mining District. In 1936 two brothers William Edward and Francis L. Blanchard along with E.L. King located the Oscura Load (Socorro County Records Mining Deeds book #112, p. 487-488) and later located the unpatented Halstead, Louise, Eva, Prairie Springs, and Calcite claims (Socorro County Records Mining Deeds book #120; Rothrock et al., 1946). Francis L. Blanchard's widow, Mrs. Ora Wallace Blanchard (1896-1968) lived a reclusive life in a small house at the base of the Oscura Mountains after her husband's passing. Visitors to the district in the 1950's and 60's were frequently guided by Ora around the family claims (R. Eveleth personal commun., 2009). From 1947 to 1954 the Blanchard claims were leased to the Portales Mining Company who expanded the Portales adit and built both a crusher and ore bin just to the north of the entrance to the adit (DeMark, 2003). They ceased working the deposit when in 1954 their ore mill, located in San Antonio, NM was raised by fire.

The year 1947 was also when claims on the Mex-Tex and Royal Flush Mine groups were established by A. R. Hickey, W. Hickey and F. Kay of Artisia, NM (DeMark and Massis, 1999). Development work on the Mex-Tex and Royal Flush claims, under the control of the Mex-Tex Mining Company, began in earnest in 1949. A 200 ton per day barite mill was built the following year. Between 1951 and 1960: the Hansonburg mining district was New Mexico's largest producer of barite with 34,868 tons or 96.5% of New Mexico's total production (Allmendinger, 1975). Most of this undoubtedly came from the Mex-Tex and Royal Flush Mine groups. Kottlowski (1953) indicates that the Mex-Tex Mining Company also had "four claims in the northern end of the Hansonburg Hills three miles west of the Oscura mountains". These are small hills in the Jornada del Muerto, on the northern end of which several small occurrences of barite-fluorite-galena mineralization are found. The Southern end of the Hansonburg Hill's is host to the copper deposit described by Eveleth and Lueth (2009). These Hansonburg Hill's are a different location than the Hansonburg Hill on which the Blanchard Mine Group is located. Mining activity by the Mex-Tex Mining Company ceased in the early 1960's, but was immediately followed by Galbar Inc., who mined and explored the Mex-Tex and Royal Flush Mine groups and shipped several carloads of lead concentrates to the ASARCO smelter in El Paso, Texas (Kottlowski and Steensma, 1979). Ownership of these claims from 1964 to 1987 is unclear (DeMark and Massis, 1999). In May of 1987, after several years of being open, the Royal Flush and Mex-Tex Mine groups were reclaimed by several individuals interested

in operating the mines for specimen recovery. Today the Upper Mex-Tex Mine group and the Royal Flush Mine are claimed by Tom Massis of Albuquerque, NM. Claims on the Ora Mine (made June 14th 1994 as the TEAA #2) and the Desert Rose Mine are currently held by Edward and Amber DeLuca and Alison Nelson respectively.

From 1958 to 1960 the Blanchard property was leased by the Sunshine Mining Company who opened six exploratory adits named the Sunshine #1 - #6, in order of chronological development. Although a total of about 701 m of tunnel were excavated no ore was ever marketed by the company (DeMark, 2003). The Sunshine mines, however, have resulted in large numbers of mineral specimens that grace collections worldwide. Little significant mining occurred in the Blanchard Mine group during the 1960's, however, the Sunshine Mining Company continued sporadic mining and exploration until 1966. This was also a period of active mineral collecting and visits with Ora Blanchard. The next company to try its luck in the district was Basic Earth Science Systems, Inc. (BESSI) who leased the Blanchard property and began several years of exploratory work in 1972. The Blanchard family relinquished all remaining interests to the claims in 1977 when Annie Lathrop, Ora's Blanchard's daughter, passed the deeds onto BESSI (Socorro County Courthouse, Mining Deeds, 1977). The year 1977 also marked the last big effort to operate the Hansonburg Mining district; this time by Hansonburg Mines, Inc. who began an extensive program of drilling exploration that year (Kottlowshi and Steensma, 1979). They also built a 400 ton per day mill complex, the foundation of which is readily visible at the North West end of Hansonburg Hill. Unfortunately, the process for separating the barite, galena and fluorite, for which the mill was constructed, did not work and as with the Sunshine Mining Company venture, no ore was ever marketed. Subsequently, control of the Blanchard Mine group went to Western General Resources until 1983. From that year until 1987 the entire Hansonburg Mining District was open to claim (DeMark and Massis, 1999) when the Blanchard Mine group was restaked by Ray DeMark and Brian Huntsman (R. DeMark personal commun., 2009). Today these claims are held by DeMark, Huntsman and Mike Sanders and the mines are also operated for mineral specimen production.

GEOLOGY

The Sierra Oscura Mountains are composed of a series of north-south trending fault blocks that are tilted to the east, with a dip of around 7°, creating a steep fault scarp on their west and a sloping highland to the east (Fig. 3). The Sierra Oscura Mountains are separated from the San Andreas Mountains to their south by a transverse shear zone that lies at the intersection of the Rio Grande rift and the Santa Rita lineament (Chapin, 1979). Displacement along the Oscura fault (Fig. 4) is the result of Rio Grande Rift extension, with major movement beginning about 7 Ma ago (Chapin, 1979). The throw along the fault increases toward the south from the Hansonburg district, and northward the fault disappears into the Oscura anticline. The Oscura fault lies about 1.5 km west of the western escarpment of the Oscura

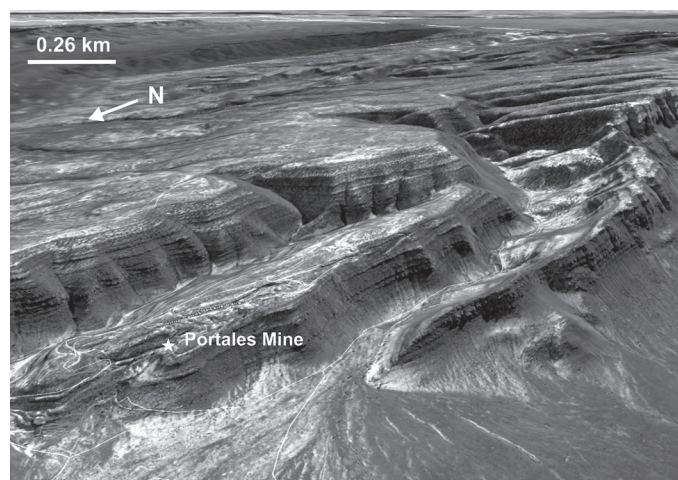


FIGURE 3. Satellite image of South Central New Mexico and the Hansonburg Mining District. Exposed strata in the Sierra Oscura dip to the east at about 7°. Image © DigitalGlobe © 2009 Tele Atlas Image NMRGIS.

Mountains, strikes N. 15° E., and is mostly hidden by alluvium (Kottlowski and Steensma, 1979). A second major fault zone (unnamed) occurs at the base of the Sierra Oscura escarpment (Fig. 4). This fault zone trends north-south, and at intervals of roughly 580 m is interrupted and progressively offset to the east, or curves to the east before dying out northward. Below the Mex-Tex Mine, this fault has a throw of approximately 168 m (Kottlowski and Steensma, 1979). Southeast-trending, high angle faults crosscut the Oscura Mountains throughout the district. These faults show less vertical offset and at the Mex-Tex Mine show a reverse-normal displacement (Putnam et al., 1983). Small faults of this type were observed adjacent to most of the mines in the district (Bosze et al., in press).

Proterozoic granites and gneisses comprise the basement rocks of the Sierra Oscura Mountains (Fig. 5). These are overlain by Pennsylvanian formations consisting of marine limestone and shale, inter-bedded with arkosic sandstone. The lower beds of the Pennsylvanian are in the Sandia Formation. In the northern part of the Oscura Mountains, the formation consists of arenaceous limestone, green to black shale, and greenish sandstone, and is 16.5-42.5 ft thick. The Madera Limestone is conformable on the Sandia Formation, and is 721-1,100 ft thick in the Oscura Mountains (Kottlowski and Steensma, 1979). Stratigraphically overlying these are Permian (Abo Formation) and Triassic red-beds and Cretaceous marine sandstone and shale, although most have been removed by erosion. Mineral deposits principally occur in the Council Spring member of the Pennsylvanian Madera group limestone. Minor vein fillings occur in Permian sedimentary rock as well as Proterozoic basement rocks such as the granites exposed in the lower portions of the Sierra Oscura Mountains. The Council Spring member is massive, banded limestone that contains solution cavities up to 50 meters in width. Some field evidence indicates that these karst features are the result of weathering during the Pennsylvanian (Putnam et al., 1983) although Lueth et al., (2005) noted features of acid speleogenesis that were contem-

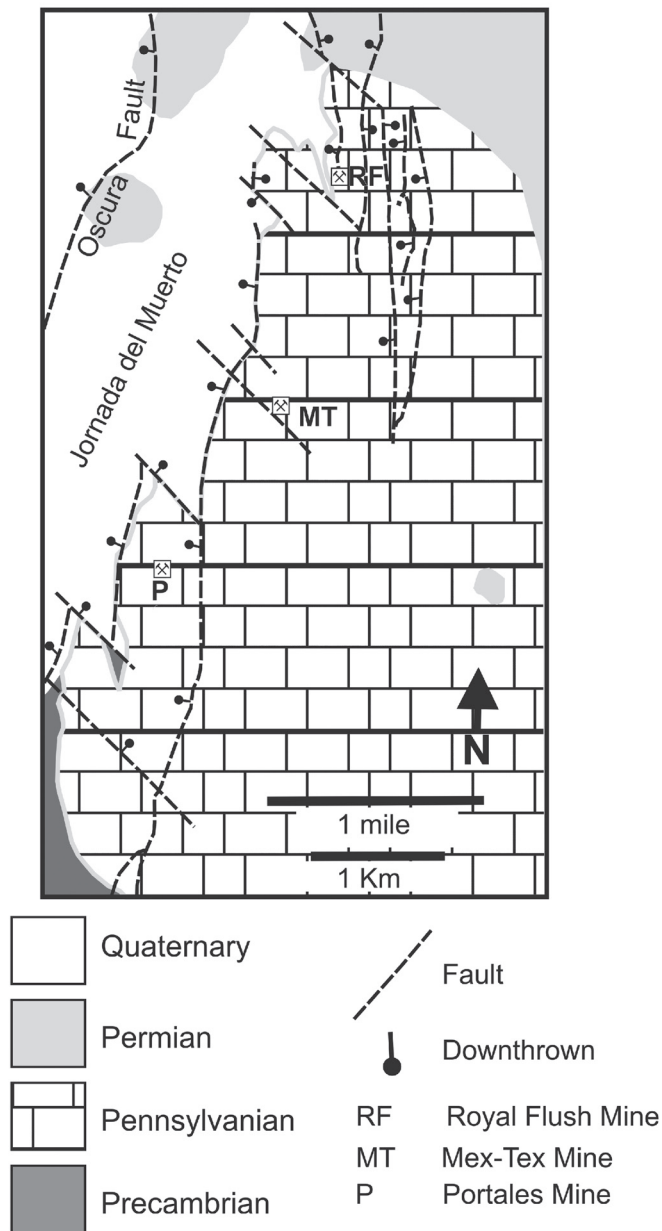


FIGURE 4. Generalized geologic map of the Hansonburg Mining District. Modified from Putnam et al. (1983), figure. 2.

porary with mineralization. This unit forms the distinctive cliff along the mountain front at the level of the mines (Kottowski and Steensma, 1979). A few sills and dikes of mid-Tertiary aged monzonite-diorite are intruded into the Pennsylvanian and Permian beds. Quaternary alluvium fills the valleys and blankets most of the surface of the Jornada del Muerto to the west.

Rio Grande Rift Barite-Fluorite-Galena Deposits

The Hansonburg Mining District is the largest of a group of roughly 30 hydrothermal deposits that have been called Rio

Grand Rift fluorite-barite-galena deposits (Fig. 6), and which are similar in many ways to MVT deposits (McLemore and Lueth, 1996; Sutphin, 1987; McLemore et al., 1998; Partey et al. 2009). McLemore et al. (1998) give a detailed review of the characteristics of RGR deposits including the Hansonburg Mining District. The RGR and MVT deposits are similar in terms of ore texture and mineralogy, mineralizing fluid chemistry, host rock lithology and genetic relationship to sedimentary basins. However,

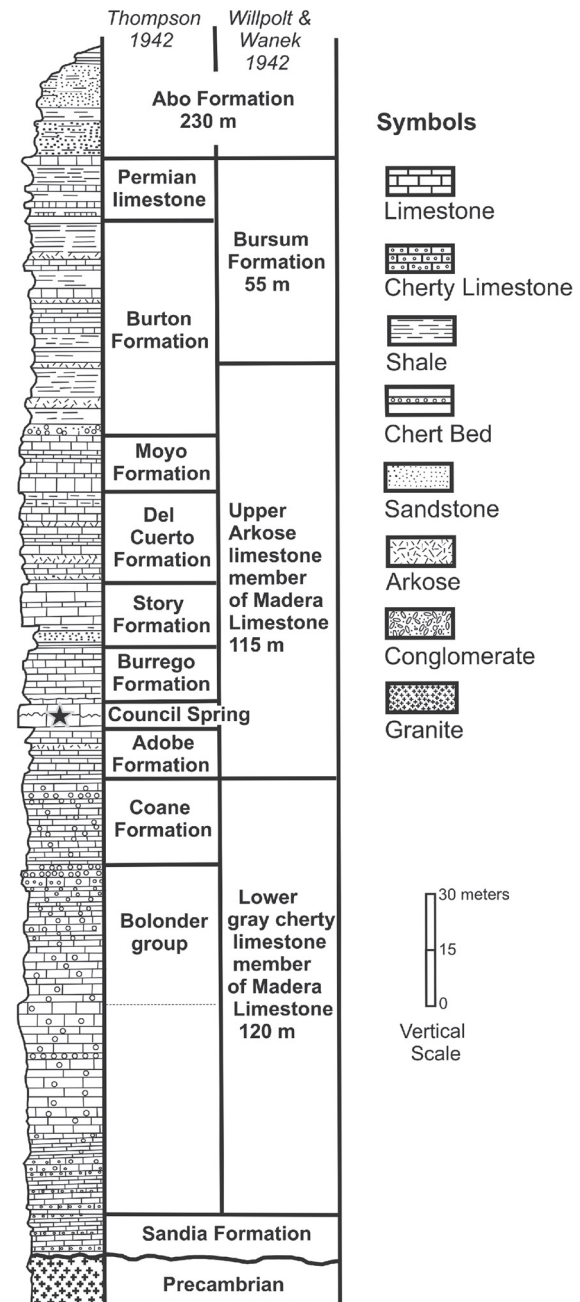


Figure 5. A Stratigraphic column of the Sierra Oscura Mountains, Hansonburg Mining District. Formation names from Thompson (1942) and Willpolt and Wanek (1951). Modified from Kottowski (1953). The star indicates the dominant host rocks for the fluorite mineralization.

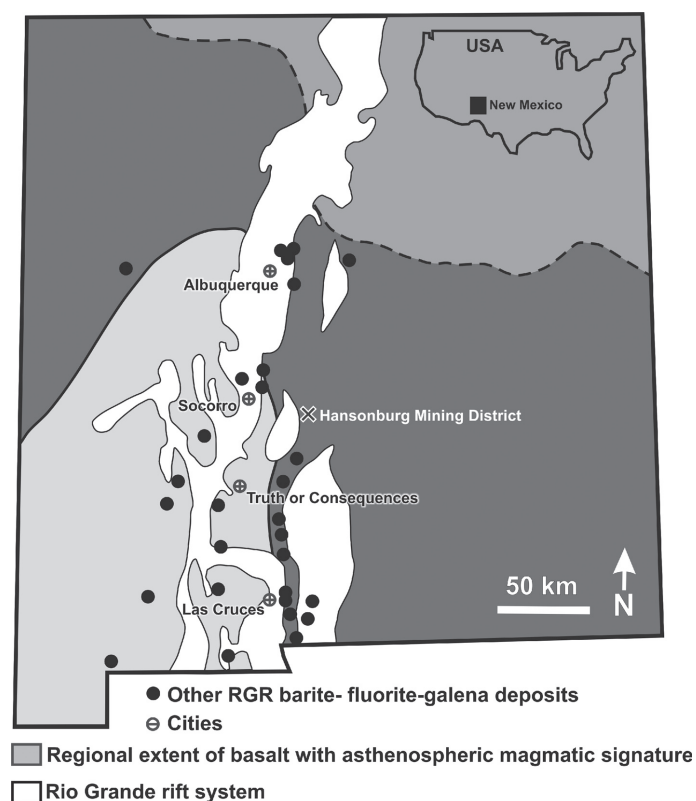


Figure 6. Rio Grande Rift (RGR)-type deposits and districts in southern New Mexico, black circles, including the Hansonburg Mining District, X (modified after McLemore et al., 1998). The distribution of asthenosphere-derived melts involved in Rio Grande Rift magmatism (McMillan, 1998) and Permian evaporites are limited to the area south of the bold dashed line.

major differences include the presence of fluorite as the predominant mineral phase, highly variable fluid inclusion salinities, the nature of their plumbing systems, and association with alkaline magmatism and extensional tectonics. Because of these differences McLemore et al. (1998) suggested that they not be considered MVT deposits.

One of the controversial aspects of RGR deposits is the origin of the large volumes of fluorine necessary to form the significant fluorite mineralization. In many of the RGR deposits fluorite is the most abundant mineral. Van Alstine (1976) suggested three possible sources of the fluorine including volatiles emanating from crystallizing alkaline magma, remelting of late fluorine-rich alkalic fractions of underlying intrusive masses, and melting of ultramafic mantle rocks bearing fluorapatite, hornblende and other fluorine-rich minerals. McLemore et al. (1998) suggested that another potential source of the fluorine could be fluorite and apatite in the Proterozoic basement rocks that underlie host carbonates and in some cases host RGR deposits as vein emplacements. Worl (1974) suggested an asthenospheric magmatic source for the fluorine, perhaps from volatiles that were released during alkaline magmatism in the region. Lueth et al. (2005) suggested a link between fluorite deposits and magmas sourced in the asthenosphere by McMillan (1998). Parthey et al. (2009) tested the asthenospheric magma hypothesis with the use of Cl isotopes as

a proxy for F, and show the potential for a significant contribution from such a source.

Paragenetic Sequence and Ore Emplacement

Mineralization in the district formed from hydrothermal waters emanating from deep within the Jornada del Muerto Basin underlying basement rocks to the west of the Hansonburg district. The high heat-flow in the Rio Grande rift (Reiter et al. 1975) and the depths of 5 to 6 km in the rift basins (Cape et al., 1983; Mitchell and Jiracek, 1983) indicate that obtaining 200+ °C waters from these basins is not problematic. Norman et al. (1985) contend that the hydrothermal brines were expelled from the basin by a combination of overburden pressure and increasing temperature. An alternative hypothesis is that basin fluids underwent topography-driven (gravity) flow (Witcher, 1988; McLemore et al., 1998). Details of the two models have are described by McLemore et al. (1998). Solutions moved laterally to the margins of the rift basin and upward along rift-bounding faults, then migrated into the little-disturbed Paleozoic sediments. F-Ca-Ba-S-Pb enriched brines were being expelled from the basin they moved into open-spaces such as bedding planes, faults, fractures and solution cavities where precipitation of barite, fluorite, galena and other minerals occurred as solutions became supersaturated.

Mineralization in the district is dominantly found within the Council Spring limestone and in the lower part of the Burrego Formation (Fig. 5). Ore mineralization has also been found in the upper massive limestones of the Coane Formation, the upper massive limestones of the Story Formation, and in the Moya Formation (Kottlowski, 1953). Massive beds of non-cherty limestone appear to be the most favorable hosts of ore because they tend to be intensely broken and shattered along faults, leading to higher porosity, whereas less massive beds break cleanly or are folded (Kottlowski, 1953). Ore bodies are principally found within fracture zones where massive limestone beds are broken and shattered by faulting and in solution cavities (paleokarst). Mineralization is massive in major cavities, with the development of many well-formed crystals. Although most mineralization is space filling, not all the open spaces in the deposit are filled; meter-size vugs occur in mineralized areas, and some karst-collapse features are essentially unmineralized (Norman et al., 1985; Taggart et al., 1989). It also occurs as banded or “coontail” ore, as described by Roedder et al. (1968), where bedding-plane cavities were filled, and which display similarities to the manto-like deposits of the southern Illinois fluorspar district. Although Roedder et al. (1968) report “extensive replacement of the limestones by crystalline and jasperoid quartz and by barite and fluorite”, most other authors report only local replacement of limestone by mineralization and indicate that it is dominantly found within preexisting porosity, i.e. along bedding planes, in fault breccia, and in solution cavities (Kottlowski, 1953; Kottlowski and Steensma, 1979; Putnam et al., 1983; Norman et al., 1985). Kottlowski and Steensma (1979) describe the banding as layers of silicified limestone, typically 2 to 11 cm thick, bounded on both sides by a mesh of galena and fluorite. Spaces in-between these bands are commonly 5 to 60 cm wide and are mostly filled by tabular barite crystals, galena

and fluorite. These bands are usually adjacent to faults, and most commonly occur on their eastward or updip side (Roedder et al., 1968).

The paragenetic sequence of mineralization started with hydrothermal solutions depositing silica, which along fracture zones partly replaced the wall rock and the sheets and blocks of brecciated limestone (Kottowski, 1953). Lasky (1932) suggested that silica partly sealed off the wall rock and prevented replacement of the limestone by ore minerals. Based on Sr isotope analyses Partey et al. (2009) suggest that much of the Ca in the fluorite is derived from a granitic source, indicating limited dissolution of host limestone. Dolomitization of the limestone is coeval in places with the first stage of silica emplacement. Silicification and dolomitization decrease in intensity away from the barite-fluorite-galena deposits. They extend approximately 300 meters away from the deposits in the horizontal, while they vertically encompasses all exposed Paleozoic rocks above and below mineralization (Norman et al., 1985). Conodont geothermometry (Cook, 1986) suggests that during this stage of mineralization the host sediments were heated to 110-140 °C within 300 m of the deposits. Siderite was also deposited in fractures and cavities, but it is not clear if this occurred during or after silicification. Almost all the faults are marked by a narrow (5-100 cm) silicified or mineralized zone associated with an irregular, wider brecciated zone 1.5 to 15 m wide. (Kottowski and Steensma, 1979).

The second stage of mineralization involved precipitation of the main ore minerals. First, there was minor replacement of carbonate by galena and barite, then open-space filling by galena, minor pyrite and sphalerite. This was followed by barite, fluorite and quartz; which locally coprecipitated (Norman et al., 1985). Based on fluorite color, morphology and fluid inclusion homogenization temperatures and salinities Roeder et al. (1968) proposed five different stages of fluorite mineralization. Norman et al. (1985) was unable to recognize these five stages when looking at the district in its entirety. Bosze et al. (in press) and Partey et al. (2009) suggest that at least two stages of fluorite mineralization are recognizable based in REE geochemistry and fluid inclusion Cl isotopes respectively. Norman et al. (1985) propose that the Pb was carried in solution as Cl-complexes. Isotope data indicate that the lead at Hansonburg is largely radiogenic (i.e., J-type lead), and suggest a source-age compatible with basement rocks (Slawson & Austin 1960; Austin & Slawson, 1961). Ewing (1979) has suggested the basement as a source of the lead at Hansonburg, whereas Beane (1974) favored Permian arkoses derived from the basement as a source.

Taggart et al. (1989) describe the third stage of paragenesis as the oxidation stage. This stage includes the formation of Ca- and many secondary Pb-, Cu- and Zn- carbonates and sulfates (Table 1). They propose a temporal and genetic overlap of the formation of primary and oxidation stage minerals as the hydrothermal system evolved from reducing, hotter conditions to cooler more oxidizing conditions. Others (Kopicki, 1962; Roeder et al., 1968; Allmendinger, 1975) have interpreted this stage to be the result of supergene oxidation of the primary ore minerals with neither a temporal or genetic relationship to the ore forming fluids. Because this stage of mineralization is volumetrically minor

TABLE 1. Minerals confirmed from the Hansonburg Mining District. Sources: Taggart et al. (1989), DeMark and Massis (1999).

Anglesite	Cyanotrichite	Plumbogummite
Antlerite	Dickite	Plumbojarosite*
Atacamite	Fluorapatite	'Psilomelane'
Aurichalcite	Fluorite	Pyrite
Azurite	Fraipontite	Pyromorphite
Baryte	Galena	Quartz
Brochantite	Goethite	Rosasite
Calcite	Gypsum	Scrutinyite**
Caledonite	Hematite	Smithsonite
Cerussite	Hemimorphite	Spangolite
Chalcopyrite	Hydrozincite	Sphalerite
Chrysocolla	Jarosite	Tsumebite
Cinnabar	Lepidocrocite	Turquoise
Copper	Libethenite	Wulfenite
Corkite	Linarite	
Coronadite	Malachite	
Covellite	Mottramite	
Crandallite	Murdochite	
Creedite	Opal (Var. Hyalite)	
Cuprite	Plattnerite	

* Identified by XRD V. Lueth personal commun. 2009.

** Type location.

(although widespread) it has not seen as much attention as the first two stages in most studies. Although Taggart et al., (1989) consider calcite to be part of the secondary or oxidation stage minerals, Roeder et al., (1968) suggest that calcite is probably part of the hypogene mineral assemblage and was the last in that sequence to be deposited.

Taggart et al., (1989) also propose that the last stage of mineralization in their evolving fluid model is very oxidizing, low temperature cave environment, pointing to speleothems that have formed over hypogene and secondary in cavities in the Sunshine #2 mine. In many of the karst type openings large masses of granular gypsum are found. This gypsum has primary fluid inclusions with no vapor bubbles suggesting entrapment at temperatures close to ambient surface temperatures (Roeder et al., 1968).

Geochemistry

The barite-fluorite-galena deposit in the Hansonburg Mining District has been extensively studied by means of petrography, paragenesis, trace element geochemistry, mineral equilibria and fluid inclusion micro-thermometry, chemical and isotopic analyses in order to understand the physical and chemical processes that formed the deposits. Fluid inclusion studies of barite, fluorite, and galena determined that the vast majority are primary and pseudo-secondary (Ames, 1958; Roedder et al., 1968; Putnam et al., 1980; Putnam et al., 1983; Norman et al., 1985; North and Tuff, 1986; Hill et al., 2000; Partey et al., 2009). Recognizable planes

of pseudo-secondary inclusions yield data that are essentially identical with coeval primaries, but the few planes of presumed secondary inclusions do not (Roedder et al., 1968). These studies indicate precipitation at relatively low temperatures, between 120 and 210°, from solutions of intermediate salinities (9–20 eq. wt.% NaCl). Liquid hydrocarbons were reported in some fluid inclusions. Ore solutions were Na-dominant brines transporting between 800 and 10 ppm Pb, with Fe, Cu, and Zn concentrations ranging from 10 to 2000 ppm, and remaining fairly consistent throughout the paragenesis (Putnam et al., 1980; Norman et al., 1985). These fluids exhibit $\delta^{18}\text{O}$ and δD values from 0.2 to –14.6 and –43 to –112 respectively (Hill et al., 2000).

Early mineralization is characterized by higher homogenization temperatures, while the later mineralization exhibits a cooling trend. No differences in salinity are exhibited from early to late mineral phases, though a slight increase is noted in a southerly direction through the district (Roedder et al., 1968; Putnam et al., 1980; Norman et al., 1985). The inclusion waters in galena reflect increased concentrations in the Ca, K and Cl for the early mineralizing solutions. These early solutions are otherwise not significantly different from the later, somewhat cooler solutions. Calculations of filling temperatures from Na/K ratios are in excellent agreement with the measured homogenization values. Salinity calculations (eq. wt.% NaCl) from analytical data reflect similar agreement with the depressed inclusion melting temperatures. Volatiles from these fluid inclusions are dominantly H_2O with about 1 mole % of the gasses CO_2 , C_nH_n ($\text{C}_1\text{--C}_6$) organic compounds, and N_2 (Norman et al., 1985). They found H_2S only in samples from early in the paragenetic sequence, when major sulfide deposition occurred. From these measured volatile concentrations and microthermometry, calculated gas-fugacity's indicate that mineralization occurred at a pressure of about 150 to 200 bars (at depths of about 1.5 to 2 km) and suggest a possible but limited unmixing of solutions; with N_2 being the primary the vapor phase.

Calculated pH values range from 4.3 to 5.1 and indicate little change in pH during mineralization. Calculated values of $f(\text{O}_2)$ and $f(\text{S}_2)$ indicate that the ore solutions became less reducing during mineralization. Table 2, from Norman et al. (1985) summarizes many of the chemical conditions of the Hansonburg fluids. They found that the principal difference between early and late mineralizing fluids was a decrease in $a(\text{H}_2\text{S})$ and related decrease in $f(\text{S}_2)$, and increase in $f(\text{O}_2)$. The data indicate that the concentration of reduced sulfur in ore fluids controlled deposition of sulfides, and that early mineralizing fluids that deposited sulfides contained approximately $10^{-2.7}$ moles H_2S .

Both temperatures of fluid-inclusion homogenization and the deposition of silica (which is highly temperature dependent) indicate that fluids were cooling during mineralization. Norman et al. (1985) concluded that the most important factor in mineral deposition was temperature decrease, that there is no indication of mixing of two fluids either in fluid-inclusion microthermometry or inclusion analytical data, and that a single hydrothermal fluid transporting both Pb and H_2S was responsible for mineralization at Hansonburg.

Studies of chlorine isotopes and Br/Cl ratios by Partey et al. (2009) from fluorite fluid inclusions from eight Rio Grande Rift

TABLE 2. Some of the physiochemical parameters of the mineralizing fluids from the Hansonburg Mining District. Table 3 from Norman et al. (1985).

	Galena Stage	Fluorite-Barite Stage
T (oC)	200	150
Ionic Strength	2.5	2.0
Ca (ppm)	3400	3400
Pb (ppm)	380	460
Cl (ppm)	45,000	2600
log $a(\text{SO}_4)$	-5.6	-4.7
pH	4.85	4.6
log $f(\text{O}_2)$	-39	-41 to -44
log $f(\text{S}_2)$	-10.3	-15 to -23
log $f(\text{CO}_2)$	1.53	1.57
log $a(\text{H}_2\text{S})$	-2.7	-5.5 to -11

Quantities averaged from measurements and calculations reported in Putnam (1980) and Norman et al. (1985). Values in parentheses represent the calculated upper and lower limits in cases where direct calculation of the quantity is not possible.

deposits, including four mines from the Hansonburg Mining District, indicate mixing of Cl from asthenospheric magmatic and evaporite sources. Based on the premise that chlorine and fluorine exhibit chemically similar behavior, and therefore are likely to be derived from the same source if chlorine is associated with rift-related magmatism, they suggest that fluorine associated with the Hansonburg Mining District is also derived from degassing of asthenospheric magmatic magmas. High $\delta^{37}\text{Cl}$ data from several deposits allows for significant involvement of an asthenospheric source for the Cl and F. They calculated the minimum possible range of Cl percentages derived from an asthenospheric source for samples from the Mex-Tex and Sunshine #2 Mines and found that as much as 49 percent could be asthenospheric in origin.

In addition, Sr and Nd isotopes were also measured from fluorites, regionally related granites, carbonates, and asthenospheric basalts to aid in understanding the petrogenesis of Rio Grande rift deposits (Partey et al., 2009). Their results show that the fluorites have Sr (and Ca by proxy) and Nd isotope ratios that are distinctly different from the local basalts and Pennsylvanian limestones but similar to the radiogenic granites, indicating that these elements are largely from a granitic source with some influence from a carbonate and/or asthenospheric source.

The $\delta\text{D}_{\text{H}_2\text{O}}$ and $\delta^{18}\text{O}_{\text{H}_2\text{O}}$ data from parent jarosite fluids suggest mixing from local meteoric water and evolved brine from deep in the basin, with a possible deep-seated magmatic gas (HF) component (Lueth et al., 2005). Jarosite $\delta^{34}\text{S}$ values which, overlap with values for barite indicate that the sulfur that formed the barite may have been derived from the dissolution of Permian evaporite by deeply circulating meteoric water (Lueth et al., 2005). The requisite sulfuric acid for the jarosite was produced by the atmospheric oxidation of H_2S that separated from hydrothermal fluids derived from the basin. The H_2S was probably derived in the basin largely as "sour gas" from the thermochemical reduction of Permian sulfate (Lueth et al., 2005). The H_2S had exception-

ally low $\delta^{34}\text{S}$, probably as a result of exchange at temperatures of ~ 150 to 200°C with the more abundant Permian ($\delta^{34}\text{S}=12\pm 2\%$) SO_4^{2-} (Lueth et al., 2005).

Fluid inclusion and geochemical analysis of fluorite suggest that the mineralizing fluids responsible for forming the RGR deposits originated as relatively dilute meteoric waters (Macer, 1978; Hill, 1994; Hill et al., 2000; Lueth et al., 2005). The meteoric waters were buried or migrated to the west of the Hansonburg district and into underlying basement rocks (McLemore et al., 1998). The fluids acquired the majority of their solutes (Na^+ , Cl^- , Ca^{2+} , Ba^{2+} , Pb^{2+} , Si , etc.), becoming Na-dominant brines, through water-rock interaction with various basin rocks such as arkosic sediments, evaporates, and Precambrian basement rocks (Beane, 1975; Bohlke and Irwin, 1992; Ewing, 1979; McLemore et al., 1998; Slawson and Austin, 1960). Parthey et al. (2009) and Lueth et al. (2005) suggest that in addition to water-rock interactions with basin sediments and basement rocks, Cl and F were also added from an asthenospheric source, most likely during magma degassing. Allmendinger (1974) suggested, based on sulfur isotopes and geochemical analysis of fluid inclusions, that the fluids leached their sulfur from the Permian evaporites. However, Lueth et al. (2005) suggest that the sulfur was alternatively derived from sour gas (H_2S) evolved from the basins based on the low $\delta^{34}\text{S}$ values they measured in the sulfides and hydrothermal jarosites. Sr , Nd (Parthey et al., 2009) and Pb (Beane, 1975; Ewing, 1979; Slawson and Austin, 1960) isotopes indicate significant contribution of these elements to solution through interactions with specifically basement granites

Geochronology

Recent $^{40}\text{Ar}/^{39}\text{Ar}$ age dating of hydrothermal jarosite coexisting with fluorite and field relationships suggest that RGR deposits formed during the last 8 Ma (from 0.47 Ma for San Diego Mountain to 7.94 Ma for the Potrillo Mountains), coincident with the later stages of rifting in central New Mexico (Lueth et al., 1999, 2004, 2005). The distribution of jarosite ages in the southern Rio Grande Rift reveals a rudimentary spectrum of ages that roughly predict rift development from south to north. The southernmost portion of New Mexico (Potrillo, Bishop Cap-Heibert mine) had the greatest number of old ages (>7.5 Ma), with one exception, the Sunshine No. 6 mine of the Hansonburg Mining District (Lueth et al., 2005). They also reported that there is a spectrum of ages for jarosite along transects across the rift. Younger ages toward the rift axis and progressively older ages are nearer the margins of the rift. (e.g., Bishop Cap district to Tonuco Mountain district in New Mexico). Even though a spectrum of ages is observed from the deposits throughout the Rio Grande Rift-type deposits, a number of geographically separated deposits share similar ages. The oldest shared age, about 8.0 Ma, occurs at Hansonburg and the deposits at the Potrillo Mountains districts. Another cluster of ages in the RGR-type deposits corresponds to mineralization from 6.5 to 6.1 Ma noted at Hansonburg (Royal Flush, Snake Pit, Portales, and Sunshine Tunnels). The mineralization during this time period spans the entire north-south geographic range (4 km) of the Hansonburg district. For a thorough review of the ages of

the RGR-type deposits refer to Lueth et al., 2005, table 1 and for Hansonburg in particular see Lueth (2009).

MINERALOGY

The most common primary (hypogene) minerals in the Hansonburg Mining District are fluorite, barite, galena, and quartz, with minor amounts of other sulfide minerals (sphalerite, chalcopyrite and pyrite). Ore minerals, for which the district was mined through out its history, include galena, barite and fluorite. Superimposed on these primary minerals are minor secondary minerals including Ca- and numerous Pb-, Cu- and Zn-carbonates and sulphates. Table 1 lists all of the minerals found in the barite-galena-fluorite deposits of the district that have been confirmed by chemical or structural analyses. Other minerals have been reported but are not listed here because of insufficient data for verification. Over the last fifty years spectacular mineral specimens have been produced from the Royal Flush, Mex-Tex, and Blanchard Mine groups. Surprisingly, very few specimens predating 1960 are known (DeMark 2003). Large crystals (>1 cm) of fluorite (up to 7 cm on edge), galena (up to 10 cm), barite, calcite, quartz, linarite, malachite and cerussite (pseudomorphs after linarite), and cyanotrichite have been found. The remaining minerals (Table 1) occur in 5 to <1 mm size crystals. Detailed descriptions of mineral specimens found in the district can be found in Taggart et al. (1989), DeMark and Massis (1999) and DeMark (2003). One of the finest collections of Hansonburg minerals can be found at the New Mexico Bureau of Geology and Mineral Resources Mineral Museum.

Fluorite

Fluorite from the Hansonburg Mining District exhibit a wide variety of forms including the cube $\{100\}$, hexoctahedrons $\{hkl\}$, tetrahexahedrons $\{hk0\}$, and less commonly the octahedron $\{111\}$, dodecahedron $\{110\}$ and trisoctahedrons $\{hhl\}$ (Taggart et al., 1989), with many crystals exhibiting combinations of forms (e.g. cubo-hexoctahedron; Fig. 7). Many cubic crystals display very minor hexoctahedral modifications, while others show significant development of this form at the corners of the cube. In other cases crystals with a rough cubic or hexoctahedral morphology are covered with many small cubo-hexoctahedrons resulting in rounded morphologies. Cubic crystals with frosted surfaces are common, and in some instances fluorites with highly etched and pitted surfaces are encountered. Color is also quite variable. The district is famous for its vividly colored fluorite crystals. "Bingham blue" (Plate 17) with its characteristic sensitivity to sunlight is the most common and most widely recognized color found at Bingham, however, crystals of teal, purple, green (of different shades and intensities) and colorless are also commonly found. Color zoning is also widespread. Zoning of color between concentric zones from the center of the crystal outward is most often seen. Many of the concentric zones in the fluorite are very thin (widths in microns) and abundant, reflecting continuous crystal growth during small environmental changes in the hydrothermal fluids from which they precipitated. Other concentric zones are

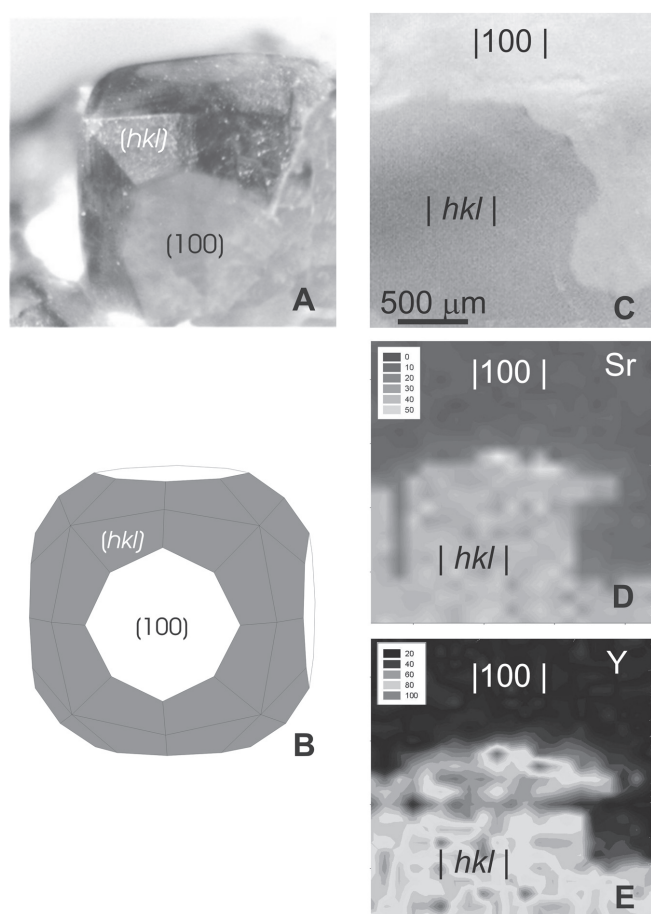


FIGURE 7. (A) Photograph of a sectorally zoned cubo-hexoctahedron from the Mex Tex Mine Hansonburg Mining District. The crystal measures 4mm across (B) Idealized schematic of the crystal in A.; colorless under the cube {100} faces and purple under the hexoctahedral {hkl} faces. (C) Plane light image of an oriented thin section parallel to (100) through a crystal similar to the one in A. The hexoctahedral sector |hkl| is purple and the cube sector |100| is colorless. (D) Plot of the concentration of Sr in ppm determined by synchrotron X-ray fluorescence (SXRF) for the area shown in C. (E) Plot of the concentration of Y in ppm determined by SXRF for the area shown in C. Modified from Bosze et al. (in press). See color plate 17.

much thicker and may represent overgrowths of one temporally distinct generation of fluorite on another. These usually have green cores (first growth stage) and purple to blue exteriors (final growth stage). It is often difficult to qualify fluorites as blue (b) or purple (p) as concentric zones often alternate between two colors (i.e., b-p-b-p). It is clear, however, that the blue and purple fluorite is paragenetically later than the green fluorite (Bosze, in press). The Mex-Tex and Royal Flush mines have also produced wonderful examples of sectoral zoning of color. In these samples symmetrically different crystal faces, and the zones beneath them (sectors), exhibit different colors. For example cubo-hexoctahedral crystals have been found with purple hexoctahedral faces, and sectors, and colorless cube faces, and sectors (Fig. 7).

Several studies have looked at the relationship between color, structure and trace element chemistry of fluorites from the district (Bosze, 2001; Wright and Rakovan 2001; Bosze and Rakovan, 2002; Wright, 2002;). The results show a complex interplay of color causing mechanisms in these samples. One mechanism that has been identified is electron transitions on simple F-centers. These are associated with the characteristic “Bingham blue” color. A more complex color center that has been identified is a REE associated fluorine vacancy. Color in the sectorally zoned crystals (Fig. 7) is the result of REE-associated fluorine vacancies.

Trace element chemistry

North and Tuff (1986) analyzed a fluorite from the Royal Flush Mine by X-ray fluorescence and found 32 ppm Mo, 134 ppm Y, 40 ppm Sr, 37 ppm Cu, 15 ppm Pb, and 102 ppm Ba. Using atomic absorption and X-ray fluorescence Hill et al. (2000) analyzed samples from the Ora, Royal Flush and Mex-Tex mines found trace elements at levels ranging from hundredths (Sc) to tens of ppm (Na, Zn, Zr, Rb, Sr, Ba, Pb and the REE).

Bosze and Rakovan (2002), using spatially resolved synchrotron X-ray fluorescence data, have shown that REE incorporation into fluorites from the district can be strongly influenced by the structure of the fluorite surface. They found that REE are differentially incorporated among structurally different crystal faces (e.g. Fig. 7) leading to sectoral zoning which in some cases was evident by color zoning. Throughout the district evidence of at least five symmetrically nonequivalent crystal growth forms were found by Bosze and Rakovan (2002) including the cube{100}, dodecahedron{110}, octahedron{111}, hexoctahedra{hkl} and the tetrahexahedra{hk0}. Each of these forms was found to have a different affinity (effective fluid-solid partition coefficient) for each of the REE measured.

Whole crystal REE concentrations, from 86 fluorite samples collected throughout the entire deposit, were determined by direct current plasma spectrometry, allowing for detailed study of intra-deposit variability (Bosze et al., in press). Total REE concentrations are typical of hydrothermal fluorites and ranged from 7 to 37 ppm. Significant intra-deposit variations in fluorite REE chemistry do exist. Based on these data and fluid inclusion Cl isotopes (Partey et al., 2009) at least two stages of fluorite mineralization are recognizable. These chemically distinct groups correlate with the paragenetic sequence of fluorite which in turn is correlated with color. Colorless and green fluorites formed first while blue and purple fluorites are paragenetically younger. Dy/La ratios are consistent with conclusions from other studies that precipitation was from a single closed system fluid that has fractionated over time (Norman et al., 1985). Alternatively, the two chemically distinct fluorite groups may have formed from two, temporally and compositionally separate fluids. This is consistent with recent geochronological and stable isotope studies (Lueth et al., 2005) that indicate the potential for at least two distinct mineralization events at in the Hansonburg mining district.

ACKNOWLEDGMENTS

We would like to thank Virgil Lueth for reviewing this manuscript and for helping to organize the NMGS 60th annual Fall field conference. We are also grateful for a review by Marc Wilson. JR would like to thank Ray DeMark, Tom Massis and Alison Nelson for their hospitality and for allowing me and my students access to their claims for research. I am also grateful to Ray DeMark and especially Bob Eveleth for many interesting conversations about the district and its colorful history.

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